Hydrothermally induced diagenesis: Evidence from shallow marine-deltaic sediments, Wilhelmøya, Svalbard

Beyene G. Hailea,*, Urszula Czarnieckab, Kelai Xic, Aleksandra Smyrak-Sikorad, Jens Jahrena, Alvar Braathena, Helge Hellevang,a,d

aDepartment of Geosciences, University of Oslo, Pb. 1047 Blindern, NO-0316, Oslo, Norway
bInstitute of Geological Sciences, Polish Academy of Sciences, Twarda 51/55, 00-818, Warszawa, Poland
cSchool of Geosciences, China University of Petroleum, Qingdao, 266580, Shandong, China
dThe University Centre in Svalbard (UNIS), Pb. 156, 9171, Longyearbyen, Norway

ABSTRACT

Sedimentary basins containing igneous intrusions within sedimentary reservoir units represent an important risk in petroleum exploration. The Upper Triassic to Lower Jurassic sediments at Wilhelmøya (Svalbard) contain reservoir heterogeneity as a result of sill emplacement and represent a unique case study to better understand the effect of magmatic intrusions on the general burial diagenesis of siliciclastic sediments. Sills develop contact metamorphic aureoles by conduction as presented in many earlier studies. However, there is significant impact of localized hydrothermal circulation systems affecting reservoir sediments at considerable distance from the sill intrusions. Dolerite sill intrusions in the studied area are of limited vertical extent (\(\leq\)12 m thick), but created localized hydrothermal convection cells affecting sediments at considerable distance (more than five times the thickness of the sill) from the intrusions. We present evidence that the sedimentary sequence can be divided into two units: (1) the bulk poorly lithified sediment with a maximum burial temperature much lower than \(60\ \text{to}\ 70\ ^\circ\text{C}\), and (2) thinner intervals outside the contact zone that have experienced hydrothermal temperatures (around \(140\ ^\circ\text{C}\)). The main diagenetic alteration associated with normal burial diagenesis is minor mechanical plastic deformation of ductile grains such as mica. Mineral grain contacts show no evidence of pressure dissolution and the vitrinite reflectance suggests a maximum temperature of \(<\ 40\ ^\circ\text{C}\). Contrary to this, part of the sediment, preferentially along calcite cemented flow bafes, show evidence of hydrothermal alteration. These hydrothermally altered sediment sections are characterized by recrystallized carbonate cemented intervals. Further, the hydrothermal solutions have resulted in localized sericitization (illitization) of feldspars, albitionization of both K-feldspar and plagioclase and the formation of fibrous illite nucleated on kaolinite. These observations suggest hydrothermal alteration at \(T > 120\ -140\ ^\circ\text{C}\) at distances considerably further away than expected from sill heat dissipation by conduction only, which commonly affect sediments about twice the thickness of the sill intrusion. We propose that carbonate-cemented sections acted as flow bafes already during the hydrothermal fluid mobility and controlled the migration pathways of the buoyant hot fluids. Significant hydrothermally induced diagenetic alterations affecting the porosity and hence reservoir quality was not noted in the noncarbonate-cemented reservoir intervals.

© 2018, China University of Geosciences (Beijing) and Peking University. Production and hosting by Elsevier B.V. This is an open access article under the CC BY-NC-ND license (http://creativecommons.org/licenses/by-nc-nd/4.0/).

1. Introduction

The quality of petroleum reservoirs depends strongly on the burial history and diagenesis, and understanding the processes that change the properties of reservoir rocks are therefore of economic importance. Diagenesis of sandstones involves physical and chemical processes that are responsible for changing the mineral
composition, texture and fluid flow properties of sedimentary rocks. Diagenesis is responsible for the formation of secondary porosity, for porosity destruction through compaction and in some cases porosity preservation to great depths (Surdam et al., 1983; Bjørlykke, 1988; Ehrenberg, 1993; Salem et al., 2000; Bloch et al., 2002). Investigating the processes and products of diagenesis is thus critical for constraining rock texture, porosity and permeability during burial (Bjørlykke et al., 1979; Worden and Morad, 2000; Ajdukiewicz and Lander, 2010; Morad et al., 2010; Taylor et al., 2010; Bjørlykke and Jahren, 2012). Moreover, understanding diagenesis improves our ability to predict reservoir quality at a local scale and modeling the evolution of sedimentary basins at regional scales (Wilson et al., 2006; Zhu et al., 2006).

As the energy demand in the world is increasing, the exploration activity for hydrocarbons moves towards more challenging systems such as basins affected by magmatic activity (Senger et al., 2014, 2017; Schofield et al., 2015b). There are numerous challenges for petroleum exploration in such basins because igneous intrusions may impact diagenesis, thermal history, seismic imaging, reservoir compartmentalization, source rock maturation, and hydrocarbon migration pathways (Holford et al., 2012, 2013; Rateau et al., 2013; Jerram, 2015; Eide et al., 2017; Grove et al., 2017; Schofield et al., 2017; Senger et al., 2017). To date, basins influenced by magmatic intrusions have become a major focus for many exploration companies. It is therefore important to understand the influence of intrusion-induced diagenesis on reservoir quality in such basins (e.g. Voring Basin, Faroe-Shetland Basin, the Northern parts of the Barents Sea, western Australian continental margin, and offshore Madagascar). This is essential to reduce exploration risk.

When sills intrude sedimentary rocks, heat can be transferred by conduction and/or by convection of fluids (Ferry and Dipple, 1991). Most published works associate the impact of sill intrusions in sedimentary strata to pyrometamorphism and/or contact metamorphism (Reeckmann et al., 1985; Karlsen et al., 1998; Mckinley et al., 2001; Grapes, 2010; Aarnes et al., 2011). It has however also been demonstrated that magmatic sills intruded into highly porous sediments have resulted in hydrothermal fluid mobility that influenced rock properties further from sill intrusions (Einsele et al., 1980), and also recent studies have shown that hydrothermally induced fluid migrations affect the temperature history and diagenesis in the host sedimentary rock (Schofield et al., 2015a; Angkasa et al., 2017; Grove et al., 2017; Senger et al., 2017).

Svalbard is suitable to examine the influence of heat-flux related to short-lived events of hydrothermal induced diagenesis in sandstones, as outcrops of highly porous Mesozoic sedimentary units that have been intruded by igneous rocks are abundant and well exposed. The Triassic (De Geerdalen Formation) and Lower Jurassic (Wilhelmøya Subgroup) sediments exposed at Wilhelmøya are time equivalent to subsurface rocks in the Barents Sea area (Mørk et al., 1999). The outcrops, in addition, provide information on the 3D spatial variability of sedimentary facies of deltaic and shallow marine strata and provide good insight into the depositional environment and the related spatial variability of diagenesis. They can also serve as an excellent natural laboratory to give insights about diagenetic alteration and ultimately reservoir quality evolution in the subsurface where sediments interact with sills. Furthermore, reservoir quality modifications as a result of diagenesis and documentation of this process is important because: (1) there are relatively few studies on hydrothermally induced diagenesis compared to normal burial diagenesis (Mckinley et al., 2001; Ahmed, 2002; Machel and Lonnee, 2002; Rossi et al., 2002; Ochoa et al., 2007; González-Acebrón et al., 2011; Grove, 2013; Holford et al., 2013; Grove et al., 2017); (2) despite several published papers on sedimentology, sequence stratigraphy, structural and tectonic evolution of the Barents Sea area (Faleide et al., 1993; Glørstad-Clark et al., 2010, 2011; Klausen and Mørk, 2014; Klausen et al., 2014, 2015; Anell et al., 2016; Lord et al., 2017), the impacts of diagenesis on reservoir quality evolution of the Triassic sequences is scarce (Mørk, 2013); and (3) Mesozoic outcrops at Svalbard are located in the uplifted parts of the Barents sea region which has a comprehensive burial history leading to different burial diagenesis compared to other basins.

On Svalbard, as a general trend, Mesozoic sandstones are well cemented and have low porosities and permeabilities. Contrary to this the Wilhelmøya Subgroup, as observed at the type location at Wilhelmøya, are composed of poorly lithified sediments with some thinner cemented intervals. These sediments have never been buried deeply (about <2 km), explaining the general lack of cementation. The diagenetic overprints occurring at the onset of the transition from mechanical to chemical compaction window makes the sequences at Wilhelmøya an excellent natural laboratory to investigate early chemical diagenesis. Moreover, by using diagenetic observations that enable placing constraints on the temperature of the sediments, the sequence offers the opportunity to shed new light on the uplift and temperature history of the area. Diagenetic processes, such as the onset of quartz cementation, pressure-dissolution at stylolites, and the transformation of smectites to illite occur in the same temperature range. Diagenetic fingerprints can therefore be used as indicators of maximum burial temperatures before uplift, and the extent of the uplift.

Temperature proxies may in addition be useful to better understand the effect of sill intrusions penetrating the sediments in this region and sediments deposited in similar settings. A larger area (~900,000 km²) of the northern and eastern Barents Sea (it covers both Norwegian and Russian territorial waters) contains abundant igneous intrusions with a volume estimate of 100,000 to 200,000 km³ predominantly in the Permian to Triassic sedimentary units across the Barents basin (Polteau et al., 2016). The sill intrusions in the study area have moderate vertical extent but have still influenced the heat budget of the area. Moreover, because sedimentary successions on Wilhelmøya have never been buried deeply, the effects of the hydrothermal alteration can easily be distinguished from the background low-temperature diagenesis. This paper examines the effect of sill intrusion-induced diagenesis compared to the normal diagenesis on sandstone reservoir properties.

2. Geological background

2.1. Svalbard geology

Svalbard is situated at the north-western parts of the Barents Shelf, bordered to the north by a rifted continental margin and to the south-west by a sheared margin (Johansson et al., 2005; Faleide et al., 2015). During the Late Cretaceous a regional uplift of the northern Barents Shelf resulted in subaerial exposure of the Svalbard archipelago (Johansen et al., 1992; Riis et al., 2008; Worsley, 2008; Dør et al., 2012, 2013; Blinova et al., 2013). The magnitude of the uplift has been suggested to be strongest in the northern and western parts of Svalbard, however, the entire Barents Shelf region has been uplifted and eroded in the late Cenozoic (Nøttvedt et al., 1988, 1992; Vorren et al., 1991). The average thickness of the strata removed due to erosion was estimated up to approximately one km in SW Barents Sea; however, the uplift was more intense north of 75° N, and with uplift close to three km on Svalbard (Nøttvedt et al., 1992; Henriksen et al., 2011; Dør et al., 2012).

Based on coal rank data, the superimposed exposed areas of the north eastern part of Svalbard have been subjected to an overburden of 2000 m (Mørk and Bjorøy, 1984). Similarly inferring to the organic geochemical analyses data, the northern Barents Sea...
the thickness of the sedimentary section removed due to erosion was almost 2000 m (Gustavsen et al., 1997). Reports based on $T_{\text{max}}$ values of $^{14}C$ and $^{14}C$ from the Olga basin area located southwest of the northern Barents Sea, indicates uplift of 1900 m in this area (Antonsen et al., 1991). In the same area, based on a regional study of vitrinite reflectance data revealed uplift of 1500–2000 m (Nyland et al., 1992). The magnitude of uplift increases northward up to 2000 m on Franz Josef Land compared to the center of the South Barents basin which experiences 400–500 m erosion (Sobolev, 2012). Using apatite fission track analyses, the northern Svalbard (Albert I Land, Ny Friesland and Nordaustlandet) has undergone...
approximately 60 °C cooling (~120–60 °C) which is equivalent to 2–3 km at a geothermal gradient of 20–30 °C/km (Dörr et al., 2012). The aforementioned exhumation and erosion burial paleo-history estimates of the areas around the Wilhelmøya Island and the vitrinite reflectivity data have indicated that the sediments at Wilhelmøya may never been subjected to high burial temperature assuming normal geothermal gradients. It should be noted that estimates from a single thermal maturity indicator likely is uncertain, however, combining the various methods may decrease this uncertainty (Henriksen et al., 2011; Haile et al., 2018).

The basin evolution of the Svalbard archipelago began with post-orogenic subsidence of pre-Devonian basement (Caledonian orogeny and older units) during extensional collapse linked to deposition of Devonian Old Red Sandstones (Johansen et al., 1992; Blomeier et al., 2003; Braathen et al., 2017). The next major tectonic event occurred late Carboniferous rifting (Gjelberg and Steel, 1981), with a transition from broad depressions to formation of fault narrow half-grabens. The Permian was a period of tectonic quiescence which is characterized by the formation of stable marine carbonate platforms (Stemmerik and Worsley, 1989). In the late Permian, a seaway connecting East Greenland and the North Sea opened, changing the marine carbonate platforms from a hot to cold-water environment. This is seen as a transition from prevalent ramp carbonates and evaporites to mainly spiculitic limestones and shales (Worsley, 2008). During the Mesozoic (Triassic), Svalbard gradually changed its position from about 45°N to approximately 60°N (Smith et al., 1994; Ditchfield, 1997). This era was characterized by sea level rise and warm climatic conditions (Hallam, 1985), with the Svalbard climate reflecting a humid temperate domain. The Mesozoic deposits on Svalbard are mostly marine, and three
main successions were deposited from Triassic to Cretaceous; the Sassendalen, Kapp Toscana and Adventdalen groups. These groups are made up of terrestrial deposits that are dissected by magmatic intrusions. During the Mesozoic period, sediments were in general siliciclastic in nature and deposited in a continental shelf setting dominated by shale and sand. There was no or little tectonic activity at the eastern and north eastern parts of Svalbard (Worsley, 2008), albeit (Anell et al., 2013) argue for some late fault activity. In contrast, the western part of the Barents Sea was a tectonically active region throughout the Mesozoic and Cenozoic times, showing comprehensive tectonic evolution characterized by several orogenic events (Faleide et al., 1984). These tectonic events affected the Barents Sea region and thus controlled the basin configurations and sedimentary responses (Gabrielsen, 1984; Mørk et al., 1989; Johansen et al., 1992; Breivik et al., 1998; Klausen et al., 2015) with the establishment of a transform fault between Svalbard and Greenland at the end of the Mesozoic. The Cenozoic saw the creation of a mountain belt in the west and a related foreland basin eastward in the central part of Svalbard. Continental to marine and back to continental sediments filled this depression, prior to uplift and erosion.

2.2. Studied stratigraphy

The samples studied here are the Upper Triassic to Jurassic Kapp Toscana Group exposures on the islands of Wilhelmøya. Wilhelmøya is an island situated in the eastern part of the Svalbard archipelago, approximately 50 km east of the main island Spitsbergen covering an area of about 120 km² (Fig. 1A). The island is located in the northeastern part of the eastern Svalbard platform near to the N–S oriented Lomfjorden Fault zone (LFZ) (Fig. 1A).

The Kapp Toscana Group includes sediments categorized into two main subdivisions: De Geerdalen Formation and Wilhelmøya Subgroup. Upper Triassic to Lower Jurassic sedimentary sequence comprises the Carnian to early Norian De Geerdalen Formation overlain by the Wilhelmøya Subgroup (Fig. 1B). The depositional setting for the De Geerdalen Formation and Wilhelmøya Subgroup has been interpreted to be nearshore deltaic environments. Based on detailed sedimentological studies, the De Geerdalen Formation is ascribed to a shallow marine to prograding deltaic sandstone succession (Mørk et al., 1999; Krajewski, 2008; Lord et al., 2017) while the Wilhelmøya Subgroup comprises a coastal plain and in deltaic to shallow marine platform environments (Worsley, 1973; Dyvik et al., 2002). The sandstones of De Geerdalen Formation are texturally and compositionally immature while that of Wilhelmøya Subgroup consists of mineralogically mature sand and sandstones (Mørk et al., 1999; Mørk, 2013). The Wilhelmøya Subgroup is further subdivided into the Flatsalen, Svenskøya and Kapp Toscana groups. These groups are made up of terrestrial deposits that are dissected by magmatic intrusions. During the Mesozoic (Maher, 2001; Nejbert et al., 2011), this activity is collectively named as the High Arctic Large Igneous Province (HALIP) (Maher, 2001). The late Mesozoic (early Cretaceous) magmatism has a widespread occurrence throughout Svalbard and is represented predominantly by sill intrusions up to 100 m thick and laterally continuous for up to 30 km. The sills studied from different parts of Svalbard contain plagioclase, clinopyroxene, alkali feldspar, Fe-Ti oxides and accessory minerals such as olivine, apatite, quartz, pyrite, chalcopyrite and bornite (Nejbert et al., 2011). Generally, similar to other islands, the sills of Wilhelmøya are tholeiitic (Veigand and Testa, 1982). The dolerite intrusions at Wilhelmøya have an undulating morphology (Fig. 2). Unlike other islands in the archipelago, sills intrusions on this island have more abundant biotite.

3. Materials and methods

Samples of Upper Triassic to Lower Jurassic aged Kapp Toscana Group sediments were collected from well-exposed outcrops at Wilhelmøya and the lowermost sequence found below the sill intrusion was logged (Fig. 2). Sampling of the DF and WS was concentrated at two locations approximately 2–3 km apart laterally. The sampling locations based on GPS coordinates are indicated in Fig. 2. The Samples collected represent the full range of stratigraphic variation within DF and WS both from tight carbonate cemented intervals and relatively uncemented intervals. No obvious weathering effects have been observed in the sandstone strata of either DF or WS. Fresh sediment samples taken directly beneath the exposed parts of sediment surface were sampled for diagenetic, petrographic, mineralogical and geochemical analyses.

Thin sections were prepared for 16 sandstone samples; 5 from the De Geerdalen Formation and 11 from the Wilhelmøya Subgroup. Conventional petrographic analyses were made to inspect the grain size, shape and mineral composition of samples. Moreover, the crystal morphology and textural relationships between the grains and pore-filling materials at thin section scale were investigated and the diagenetic features were examined using a JEOL JSM-6460LV scanning electron microscope (SEM). The SEM coupled with an energy-dispersive spectrometer (EDS) was used to perform spot chemical analyses to obtain semi-quantitative mineral compositions. Quartz cement is often difficult to delineate clearly using optical microscope and SEM coupled with backscattered electron (BSE) images. Therefore, SEM coupled with cathodoluminescence (CL) analyses were used to differentiate quartz overgrowths from detrital quartz. Moreover, stub samples were analyzed using SEM in order to complement the thin-section study and to decipher the morphology of some clay minerals where it was not obvious in the 2D thin-section (for example illite, chlorite and kaolinite).

X-ray powder diffractometry (XRD) analyses of bulk-rock samples were performed: (1) to complement optical petrography and SEM/EDS in order to identify and quantify sandstone constituents, and (2) to guide the thin section analysis. About 3.5 g of each sample were crushed, milled in ethanol in a McCrone micronizing mill and then dried at 60 °C. Randomly oriented powders were prepared by top loading into PMMA (Polymethylmethacrylate) sample holders designed with concentric circular geometry grooved shallow wells. The powder diffraction patterns were then
collected using a Bruker D8 Advanced diffractometer with Cu Kα radiation at a wavelength of 1.5406 Å. All XRD data collected were first analyzed for phase identification using the search-match module of the EVA software using the reference databases ICDD PDF-2 and COD. After phases were identified with EVA, they were further analyzed based on the Rietveld quantitative X-ray diffraction refinement approach (RQXRD). Rietveld QXRD was performed using the Profex-BGMN-bundle software version 3.5.0 (Döbelin, 2015) and using the BGMN crystal structure files.

Vitrinite reflectance, fluid inclusion, and stable carbon (δ13C) and oxygen (δ18O) isotope analyses were performed on selected samples. One coal sample, from the Lower Jurassic strata, was analyzed for vitrinite reflectance at Applied Petroleum Technology (APT) at Kjeller, Norway, and all analytical procedures followed NIGOGA, 4th Edition (Weiss et al., 2000). Reflected light studies were carried out using polished bulk samples mounted in resin blocks, and polished isolated kerogen samples, mounted in resin on petrographic slides. Reflectance measurements were made with a Zeiss Epiplan-Neofluar 40x oil immersion objective and 10x eyepieces, with an inherent tube magnification of 1.6x giving a total visual magnification of 640x. R₀ (Random) reflectance measurements were made in non-polarized light setting. All the performed measurements were calibrated with regard to standards of known reflectance. The maximum burial temperature was estimated from R₀ using:

$$T \left( ^\circ C \right) = \frac{\ln R_0 + 1.68}{0.0124}$$

(Barker and Pawlewicz, 1994). This model was chosen because it can reproduce comparable results as that of the kinetic model developed by Sweeney and Burnham (1990). The Sweeney and Burnham (1990) kinetic model (thermal history) predictions of maximum burial temperature can be used where the burial history of the geological systems is well known.

Eight outcrop samples were prepared as polished thick sections of approximately 60–70 μm thick for fluid inclusion petrographic analyses and microthermometry measurements. Microthermometric determinations on fluid inclusions were carried out on quartz overgrowths and pore-filling calcite cement. The fluid inclusion data were collected mainly on primary inclusions using a Zeiss Axioscope A1 APOL digital transmission microscope coupled with a calibrated Linkam. TH-600 heating and cooling stage at China University of Petroleum, Qingdao, China. The instrument enables measurement of temperatures of phase transition in the range of −180 °C to 500 °C. Homogenization temperature (T_h) measurements were determined using a heating rate of 10 °C/min when the temperature was lower than 60 °C and 5 °C/min when the temperature exceeded 60 °C. The measured temperature precision for T_h is ±1 °C. The salinity data were calculated from freezing point depression in the system NaCl-H2O for aqueous inclusions (Bodnar, 2003).

Stable carbon and oxygen analyses were made at Institute for Energy Technology (IFE) at Kjeller, Norway. Analyses were performed on four calcite-cemented sandstone samples: two from the De Geerdalen Formation and two from the Wilhelmøya Subgroup. These data are reported in per mil (‰) relative to the Vienna Pee Dee Belemnite (V-PDB) standard. The samples were heated to 400 °C for 4 h in order to remove any organic compounds. Approximately 100 μg sample was then transferred to a 10 mL vacutainer and put in a temperature controlled aluminum (Al) block. The sample was flushed with helium (He) gas for 5 min. Each bulk sample was reacted with 0.1 mL of concentrated phosphoric acid (H₃PO₄) at 30 °C for 2 h. The produced CO₂ gas (calcite fraction) was then flushed out with helium to a Poraplot Q GC column and analyses were done directly on a Finnigan MAT DeltaXP Isotope Figure 3. QFL composition of the studied thin sections; Q—quartz, F—feldspars, L—lithoclasts. The QFL triangle classification is after McBride (1963), Dott (1964), Folk et al. (1970), Williams et al. (1982), and Pettijohn et al. (2012). Petrographic point count data are presented in Appendix A. Table 1.
ratio Mass Spectrometer. Based on repeated analyses in the laboratory with respect to standards (V-PDB), the precision of reported results was ±0.1‰ (2σ) for both δ18O and δ13C. Since we do not know the composition of the formation water that calcite precipitated from, we have used the equation established by Keith and Weber, (1964) that linearly relate δ13C and δ18O with the Z parameter. Z-values above 120 are classified as marine while those below 120 are meteoric (fresh) water. Calcite temperatures were calculated using the calcite-water fractionation factor equation (1000ln(a_{cal-water} = 2.78 × 10^2 T−2.89)) by assuming the water isotopic composition relative to SMOW based on the Z-value computations (Friedman and O’Neil, 1977).

4. Results

Both De Geerdalen Formation (DF) and Wilhelmøya Subgroup (WS) have sediment layers (<0.5–10 m thick) that are heavily cemented by mainly carbonates. There is apparently no relation between the frequency of occurrence and stratigraphic position of these cemented units or between cement and distance to the magmatic intrusions.

4.1. Mineralogy of the DF and WS sands and sandstones

The DF and WS sediments are very fine to medium-grained and are usually moderately well sorted, but range from well to poorly sorted. Quartz, feldspars and lithic rock fragments are the most frequent framework components. Lithic rock fragments are mainly of igneous and metamorphic origin but some sedimentary fragments are also found. Micas (mostly muscovite), glauconite, chert, hornblende and bioclasts represent other detrital grains observed. The studied sandstones are mostly feldspathic litharenites to sublitharenite, but also lithic arkoses and subfeldsarenites (Fig. 3, appendix A). DF and the lower section of

![Thin section XPL photomicrographs, bulk XRD mineralogy and SEM photomicrographs DF samples collected from calcite-cemented intervals below the sill intrusion at Wilhelmøya. Ca – Calcite, DF – De Geerdalen Formation.](image)
WS (Flatsalen Formation) sediments are mineralogically immature but samples collected from the middle section of WS (likely Svenskøya Formation?) are mineralogically mature (Fig. 3). The average framework composition of this middle section is Q88F5L7.

Bulk XRD quantitative analyses suggest that quartz is the main framework mineral in all of the samples. K-feldspar, kaolinite, albite, muscovite, calcite, and Fe-rich chlorite are the remaining dominant minerals, while pyrite was found in very few of the samples and gypsum only in one sample. The XRD suggest that the amount of quartz ranges from 10% to 98%, K-feldspar from 0.6% to 7.9%, albite/plagioclase from 1.2% to 21%, kaolinite from 1% to 9.4%, chlorite from 0.7% to 11%, Muscovite/ilite from 1.3% to 13%, calcite from 1.2% to 86%, siderite from 8% to 46%, pyrite from 0.2% to 0.8%, and gypsum constituted about 1.5%. The results of the quantitative XRD Rietveld refinement, SEM and optical microscope micrographs of selected samples as a function of proximity to the sill are presented in Fig. 4. The amount of calcite cement in sampled carbonate cemented intervals is 17% at 1 m, 29% at 2 m and 33% at 50 m distance from the sill intrusion (Fig. 4).

4.2. Compaction

The sediment fabric is a result of mechanical and chemical compaction processes (Einsele, 2013). The most common textures resulting from compaction are sediment grains floating in cements, and tangential-, straight-, sutured-, and concavo-convex intergranular contacts (Fig. 5A–C). The DF and WS sediments exhibit dominantly point contacts and floating grains followed by long or line contacts, but a few concavo-convex and sutured contacts are also identified (Fig. 5A–D). Both, in the very porous weekly consolidated and calcite-cemented sedimentary units, mica grains show no evidence of plastic ductile grain deformation, but minor mica deformation is noted (Fig. 5G).
4.3. Quartz overgrowth

The samples belonging to DF and WS show quartz overgrowths on detrital quartz. Quartz cement includes quartz overgrowth (Fig. 6A–H) and microquartz (Fig. 6B). Microquartz occur as coatings on detrital framework grains in the WS sediments. Thin sections under an optical microscope reveal clearly the quartz overgrowths (Qo) on detrital quartz grains (Fig. 6A–H). The quartz overgrowths, however, appear to be rounded or dissolved or parts being spalled off (Fig. 6A, C–E). Quartz overgrowths can be distinguished readily from detrital quartz grains due to the presence of dust rims on detrital quartz grains (Fig. 6A–H). However, it is difficult to distinguish quartz overgrowths from detrital quartz in SEM-BSE micrographs and such distinction may also be misleading unless supported by other analytical methods. SEM-BSE display quartz overgrowths that appear euheiral (Fig. 6D). Close-up inspections of BSE micrographs however indicate that also these surfaces have been dissolved or abraded (Fig. 6A, C, and E). This is further supported by SEM-CI micrographs (Fig. 7A–F). Quartz overgrowths (red arrows) often have very low CI intensity or non-luminescent (appears dark) compared to the detrital quartz grains (appears bright) (Fig. 7A, D). SEM-CI micrographs furthermore reveal a high fracture intensity of detrital quartz grains and fractures that have been healed with quartz cement (Fig. 7A, D).

4.4. Feldspar alteration

Alteration of plagioclase and K-feldspar is only found in the calcite cemented sedimentary units within DF and WS. Most plagioclase grains have been altered to some degree and pervasively leached grains are most commonly observed (Fig. 8A–C). Plagioclase leaching leaves large (~30 μm to 150 μm) secondary pores with little remnant grain material (Fig. 8A–C). Plagioclase leaching results in some albite formation within the voids previously occupied by detrital plagioclase grains. The replacement of the original detrital plagioclase grains by albite has resulted in the formation of aggregates of small euhedral albite crystals (Fig. 8A–C). Due to calcite cement, the secondary porosity formed from plagioclase dissolution is a stable void maintaining the shape of the dissolved plagioclase grains (Fig. 8A–C).

The potassium feldspars appear to be generally fresh and less altered compared to plagioclase (Fig. 8B and C). Crystallization of very thin albite rims (~2 μm) are predominantly observed around K-feldspar grains, and may indicate leaching of K+ and recrystallization of the K-feldspar (Fig. 8A–C). The rims around K-feldspar grains are pure albite in composition and the albite crystals have abundant micro-intercrystalline porosity (Fig. 8D). In contrast to this, in the units lacking carbonate cement, there was no crystallization of albite rims observed around K-feldspar grains (Fig. 8E), likewise there was no plagioclase albization (Fig. 8K).

Near K-feldspar grains with albite rims, fibrous illitic type clay phases occur mainly accompanied with microquartz crystals (Fig. 8F and G). These clay minerals are composed of interwoven fibrous illite bridging the pores between microquartz grains. Illitization is difficult to observe in thin-sections. However, SEM examination of the texture of authigenic illite using stub samples, show that fibrous illite nucleated and grew onto kaolinite around altered K-feldspar grains (Fig. 8H). SEM-EDS analyses of the illitic clay phase yields the major elements: K, Al, Si and O (Fig. 8F). Similar to the albite formation, illite formation is only observed in the carbonate-cemented intervals.

Plagioclase and K-feldspar alteration into sericite (illite) was recognized in the carbonate cemented layers (Fig. 8I and J), but sericite was not found associated with plagioclase and K-feldspar in sedimentary units lacking carbonate cement (Fig. 8K and L). Relict plagioclase grains show grain contacts with poikilitopic calcite cement and relict grains are sometimes engulfed by the cement (Fig. 8A–C).

4.5. Clay minerals

Kaolinite (Kao) is found in primary pores of both carbonate cemented and uncemented sandstone layers associated with mica and K-feldspar (Fig. 9A–E). Kaolinite crystals are observed between mica flakes and at the inter-fingering edges of mica (Fig. 9A–C) and also associated with relict K-feldspars (Fig. 9D–F). Replacement of mica by kaolinite was commonly observed in mica grains squeezed between rigid framework grains (Fig. 9A).

The most common clay mineral observed in addition to kaolinite is Fe-rich chlorite. SEM inspection and XRD analyses revealed that chlorite is present in all of the studied De Geerdalen Formation and the Wilhelmøya Subgroup sandstone samples. The Fe-chlorite occur as masses of interwoven flakes of small crystals (~0.2 μm) with loose internal structure arranged in a chaotic pattern at the surface of framework grains (Fig. 9G–I), but also noted as massive aggregates (orange arrows) in the pore space (Fig. 9H). However, very close to the surface of the grains, the chlorite crystals occur with parallel or slightly oblique orientation (Fig. 9I). The chlorite coats are not continuous and display thickness variations along the surface of detrital grains (Fig. 9G and H). They predominantly occur at the embayments (yellow arrows) but are absent or scarce at rounded and flat edges of the grains (red arrows) (Fig. 9G and H). SEM stub samples display the clay coatings with an overlapping flaky aggregate and ragged outlines (Fig. 9). SEM-EDS analyses of the clay coats and rims give similar elemental composition interpreted to be primarily a chloritic-type clay in composition (Fig. 9K). There is no clear chlorite recrystallization noted (Fig. 9L).

4.6. Carbonate cements

Carbonate cements are only found sporadically in DF and WS. Some intervals, generally less than 1–2 m thick, are however heavily cemented. A closer examination of these cemented beds shows that most of the pore space is filled with calcite and siderite, and the remaining porosity is less than 4% (Fig. 10A–F). The inter-granular volume (IGV) of the cemented intervals is generally higher (around 40%) than the non-cemented intervals (Fig. 4). The sandstones are cemented by poikilotopic calcite (Figs. 4 and 10). In some parts of the sandstones, the poikilotopic calcite cement has replaced partially to pervasively the framework grains such as plagioclase (Fig. 10A, D) which resulted in the presence of oversized calcite cement. The transformation of the original carbonate cement via the dissolution of the hot focused fluids into a new and different carbonate fabric is the recrystallization we refer to backscattered SEM images. The backscattered SEM images reveal that recrystallized calcite cement is primary fill pore spaces (Fig. 10A). Similarly, recrystallized calcite cement interfingers into quartz grains (Fig. 10C). Pervasively etched calcite cements are visible under optical microscope (Fig. 10D). Moreover, stub samples inspected under SEM display a micro-topography with dissolution pits, smooth surfaces and sharp edges unequivocally supporting textures observed under optical microscope (Fig. 10E and F). Calcite crystals display a scalenohedral pyramidal geometry (Fig. 10E and F).

4.7. Paleo-temperature proxies

4.7.1. Sediment temperature from vitrinite reflectance

The vitrinite reflectivity (VR) for coal sample in the upper part of the section was 0.30% based on 37 measurements (Fig. 11A). Based
Figure 6. SEM and optical microscope cross-polarized thin section micrographs illustrating. (A) Optical micrographs of rounded quartz overgrowths (Qo) and dissolved quartz overgrowth (Qod) around detrital quartz (Qd). The sediments display well developed “dust” lines marking the boundary between the detrital quartz and overgrowth. The quartz overgrowth thickness varies from 2 to 5 μm. (B) Microquartz coating at the surface of a detrital quartz grain. (C) Closeup view of detrital quartz showing rounded and dissolved quartz overgrowth demarcated by dust rims (red arrows). (D) Quartz overgrowths that looks like euhedral crystal faces. (E) An enlargement of the area outlined by the red box in Fig. 6D reveals abrasions and breakage (discontinuous euhedral faces) quartz overgrowths or dissolved quartz overgrowths. (F) Photomicrographs of rounded quartz overgrowths
on the model of Barker and Pawlewicz (1994), this vitrinite reflectivity value corresponds to a maximum burial temperature of 38.4 °C.

4.7.2. Temperature of formation of the diagenetic phases

Aqueous inclusions for quartz overgrowths with irregular or rounded type of faces (Fig. 6) were homogenized in the range from 89.8 °C to 128.6 °C with an average temperature value of 109 °C (Fig. 11B), but these grains were most likely recycled. It was not possible to find any quartz that is without doubt authigenic with euhedral face and containing fluid inclusions. Therefore, no temperatures were therefore recorded from the authigenic quartz actually that could have formed in-situ.

Calcite cement in the heavily cemented parts contained aqueous fluid inclusions with \( T_h \) between 100 °C and 138 °C with an average value of 123 °C (Fig. 11B). The \( \delta^{18}O_{\text{VPDB}} \) values vary from −7.3‰ to −10.2‰ and \( \delta^{13}C_{\text{VPDB}} \) values range from −13.3‰ to −20.6‰ (Table 1). The Z-values were calculated based on \( \delta^{13}C \) and \( \delta^{18}O \) ranging from 97 to 105. The gray shaded region illustrates the possible ranges of precipitation temperatures (65–140 °C) assuming the waters involved were of meteoric origin based on the Z-values calculation with \( \delta^{18}O_{\text{SMOW}} \) ranging from −3.5‰ to −7.5‰ (Fig. 11C).

5. Discussion

Diagenetic signatures reveal that the DF and WS sediments have been subjected to two types of thermal conditions: (1) normal diagenesis resulting in increasing temperature as a function of increasing burial depth, and (2) hydrothermal induced diagenesis resulting in increasing temperature as a function of increasing burial depth, and (2) hydrothermal induced diagenesis. The following section discusses the relative influence of both normal and sill induced diagenesis on present reservoir quality of these sediments.

5.1. Normal diagenesis

Ductile grains such as mica register the effects of mechanical compaction (Chuhan et al., 2002). The detrital mica being predominantly flat or undeformed to slightly bent in DF and WS sediments indicate shallow burial of the sediments before uplift. This is consistent with the types of grain contacts identified in the studied sediments. The abundance of the grain contacts being tangential and long including floating grains indicates that the sediments have been subjected to little mechanical compaction (Wilson and McBride, 1988). Concavo-convex contacts being uncommon and the existence of only incipient sutured contacts between adjacent grains indicate that the sediments underwent no or very limited chemical compaction. This is consistent with the absence of euhedral quartz overgrowths. This indicates that the sediments have not reached temperatures in excess of 60–70 °C (Walderhaug, 1994). This is in accordance with the vitrinite reflectance (VR) data translating into a temperature of about 38.4 °C (Fig. 11A). The VR is one of a number of organic thermal maturation indicators that provides the maximum temperature exposure of sedimentary rocks. However, the empirically based or kinetic translation of VR values to paleo-temperature values is still challenging. Furthermore, all the diagenetic evidence such as microquartz coatings, feldspar dissolution and precipitation of kaolinite and early calcite cementation, indicate that these sediments have not reached quartz precipitation temperatures (>65 °C) before they have been uplifted. All the above-mentioned data indicate that the sediments at Wilhelmøya have been subjected to the shallow burial depth (about < 2 km).

The quartz overgrowths identified in this study do not display euhedral shape, but are rounded and show dissolution features (Fig. 6). The existence of these rounded and dissolved quartz cement suggest that the quartz cement overgrowths were not forming in-situ but rather are inherited overgrowths, i.e. redeposited grains (Sanderson, 1984). Quartz overgrowths possibly represent remnants of cement formed at the detrital quartz surface from a previous sedimentary cycle at deep burial. This is unequivocally supported by the homogenization temperature \( (T_h) \) measurement from these quartz overgrowths. \( T_h \) ranges from 90 °C to 130 °C (Fig. 11B). The quartz overgrowth \( T_h \) values undoubtedly depart from the normal burial history of the sediments in study area.

The precipitation of microquartz crystals requires fluids with high silica supersaturation which provides a large number of small nuclei rather than few and larger crystals. Such supersaturation is most commonly provided by the dissolution of unstable biogenic silica and other small silica fossils (Williams et al., 1985; Taylor et al., 2010). The most commonly cited biogenic precursor phase for the growth of microcrystalline quartz, for instance in the Upper Jurassic reservoir rocks from North Sea, is sponge spicules (Aase et al., 1996). However, spongy spicules have not been observed in these samples during optical microscope and SEM investigations. At this stage, it is not clear what was the source of microcrystalline quartz.

As noted in this study, most of the kaolinite was pore-filling while occasional kaolinite crystals were observed between mica flakes and at the inter-fingerings of mica. Authigenic kaolinite in general the alteration product of feldspars and micas at shallow burial depth related to flushing by meteoric waters either during early diagenesis or after structural inversion (Byllykke, 1980). Similarly, the authigenic kaolinite observed in DF and WS sandstones have formed as a consequence of feldspars and micas dissolution by meteoric water.

Authigenic chloride may be formed locally as a replacement of reactive grains such as volcanic rock fragments (VRF), and transformation of the precursor clay minerals such as berthierine (Aagaard et al., 2000; Haile et al., 2015). Recrystallized authigenic chloride coating form mainly from iron-rich precursor clay phase and will show a perpendicular orientation relative to the grain surface. Well-developed crystals having euhedral morphology will therefore commonly be arranged in an edge-to-face stacking pattern (Wilson and Pittman, 1977; Pittman et al., 1992; Grigsby, 2001; Haile et al., 2015). Moreover, such radial authigenic chloride coats are often thick and continuous on the detrital grain surfaces. However, in this study, the chloritic clay coats are: (i) attached tangentially at the surface of detrital framework grains, (ii) patchy and discontinuous, sparsely distributed at rounded and flat surfaces but thick at the embayments in the form of loose aggregates, and (iii) the chlorite crystals are poorly-developed. These evidences indicate the detrital nature of the origin of the chloritic-type clay coats (Wilson and Pittman, 1977; Moraes and De Ros, 1990). SEM-BSE micrograph of the stub sample show overlapping flaky aggregates with a range of outlines oriented nearly tangential to the surface of the grains (Fig. 9) is also another clear evidence suggesting the detrital origin of the clay phase covering the surface of the grains (Moraes and De Ros, 1990). Recrystallization of chlorite takes place above about 80 °C (Aagaard et al., 2000). This will in most...
cases result in growth of radial chlorite crystals on top of the tangential precursor towards the pore. This will often result in a brighter BSE greyscale image towards the pore space compared to the grain side because the recrystallized chlorite contain more Fe. Fig. 9L indicates no such chlorite recrystallization indicating that the chlorite coats never reached temperatures approaching 80 °C.

5.2. Evidence of hydrothermal induced diagenesis

Hydrothermal fluids have an effect on porosity and permeability evolution in reservoir rocks and the thermal maturation of source rocks (Karlsen et al., 1998; Ochoa et al., 2007; Holford et al., 2013; Senger et al., 2014; Grove et al., 2017). There have been discussion in the literature regarding the criteria necessary to identify ancient hydrothermal heating events based on geochemical reaction signatures (e.g. Machel and Lonner, 2002). In sedimentary sections with anomalously high paleotemperatures compared to burial history models, geochemical reactions may be used to identify the influence of hydrothermal systems (Ochoa et al., 2007). However, in this study, sill-induced hydrothermal diagenetic processes can unequivocally be separated from normal diagenesis, because: (i) The sediments at Wilhelmøya were only at shallow burial depths before uplift (Mork and Bjorøy, 1984) and (ii) the diagentic signatures studied herein and also the vitrinite reflectance data show only shallow burial processes, except for the carbonate cemented layers.

5.2.1. Diagenetic fingerprints in the carbonate cemented sedimentary units

In the carbonate cemented sedimentary units diagentic evidence, such as sericitization (illitization) of feldspars, feldspar albition, and fibrous illite formation, suggest more different diagenesis than the normal burial diagenesis. Sericitization (illitization) of feldspar grains in cemented intervals further strengthens the interpretation of local hydrothermal alteration (Meunier and Velde, 1982; Que and Allen, 1996). The replacement of feldspar by sericite (illite) occurs when hydrothermal fluid temperatures reaches above 100 °C (Verati and Jourdan, 2014).

The IGV values for the carbonate-cemented intervals are high and the presence of floating framework grains and straight flat mica grains engulfed by carbonate cement without any sign of deformation indicate an early near surface formation (before significant burial compaction) of the calcite cement. Therefore, the calcite fluid inclusion data, giving homogenization temperatures, between 100 °C and 138 °C reflect hydrothermal induced recrystallization of calcite. Calcite cements show pervasively etched micro-topography with dissolution pits and smooth surfaces with sharp edges (Fig. 10D–F). This suggests the dissolution-reprecipitation process that took place when hydrothermal fluids were focused around the carbonate-cemented units. Hellevang et al. (2017) documented calcite crystals with similar morphology grown at high temperature laboratory experimental conditions. The precipitation-dissolution processes take place as a result of competitive environment for divalent ions between clay minerals and carbonates (Hellevang et al., 2017). In the experimental study (Hellevang et al., 2017), similar to the natural setting reported herein, the newly formed calcite crystals display both etched and euhedral crystal outlines (see Fig. 2 in Hellevang et al., 2017).
The analyzed calcium carbonates are depleted in $^{13}C$ with $\delta^{13}C_{\text{PDB}}$ values ranging from $\sim -7.5\%$ to $\sim -10.2\%$. The carbon isotope values may suggest derivation of dissolved carbon either from oxidation of methane or microbial sulphate reduction. Moreover as documented in Grove et al. (2017), this carbon isotope value may suggest magmatic carbon. The carbon isotope values of this study are not definitely from oxidation of methane even though likely from bacterial sulphate reduction. Sulphate reduction reaction drives alkalinity and often produces pyrite. The majority of cemented beds lack however, correlation between the calcite cement and pyrite content, which precludes the importance of sulphate reduction. The carbon isotope compositions of calcite
samples ($\delta^{13}C$ of $-7.5_{\text{psm}}$ to $-10.2_{\text{psm}}$) are not indicative of a specific source of the bicarbonate ion. These values could result from precipitation in pore waters bicarbonate ions supplied from two sources. Alternatively, the carbon isotope composition could reflect the dissolution, equilibration and re-precipitation, of in situ carbonate cements with a single CO$_2$ isotopic composition produced during hydrothermal invasion resulting from sill intrusions (Grove et al., 2017). The Z-values for the paleowater ranged from 97 to 100 suggests that calcite precipitated from meteoric water rather than saline water, but this may be highly uncertain. The $\delta^{18}O$ values of calcite cements ($-13.3_{\text{psm}}$ to $-20.6_{\text{psm}}$) indicate precipitation at temperatures of approximately 65 °C to 140 °C (Fig. 11C). The range obtained may reflect variations in diagenetic zones (normal burial diagenesis and hydrothermal induced diagenesis) or variations in
δ₁⁸O of the original pore waters during early diagenesis due to fluctuations in dry or wet periods or the carbonates were affected by multiple thermal pulses during the emplacement of the magmatic sills. Most likely the temperature and isotopic composition range obtained in this study is a result of water percolation resulting from a sill intrusion induced hydrothermal convection cell.

The origins of the carbonate-cemented zones in the sedimentary strata are not clear, however, they may be derived from: (1) dissolution products of unstable Ca- and Mg-bearing non-carbonate minerals and (2) reprecipitation of dissolved bioclastic particles (Fig. 12A–D). The existence of layers composed mostly of bivalves (Fig. 12D) indicate that at least some of the carbonate layers may be derived from dissolution of mineralogically unstable bioclasts which are predominantly bivalves (Fig. 12A–D). Accumulated bivalve assemblages can easily be transported and redeposited in layers (Fig. 12D, the bottom part). This part of the section as shown in Fig. 12, is the conglomeratic Slottet member of the Wilhelmøya subgroup, and it contains phosphate mineral nodules in addition to bivalves. Bivalves are not, however, restricted to this bed, rather observed associated with carbonate cemented units in the sedimentary succession.

There was no sign of albitization of plagioclase and K-feldspar in sedimentary units lacking carbonate cement, however, in the calcite-cemented intervals the K-feldspar and plagioclase grains have been albited. The absence of feldspar albition in the majority of the sedimentary units except in the carbonate-cemented intervals suggest that the host sediments have not reached the feldspar albition temperature window.

Plagioclase is preferentially albited compared to K-feldspar (Morad et al., 1990), whereby albite crystals predominantly exist in the voids left by dissolved detrital plagioclase. Albite grains exist mainly as aggregates of euhedral albite crystals that replace the detrital plagioclase grain. Albite is formed mainly as very thin rims on K-feldspar grains. The replacement appear to be pseudomorphic within K-feldspar while mainly non-pseudomorphic in the plagioclase. This suggests that the mechanism of plagioclase and K-feldspar albition processes are quite different. The very thin albite rims found on K-feldspar grains may signify the low abundance of secondary potassium sinks observed. Apparently, small amounts of fibrous illite formed locally associated with kaolinite and dissolved feldspar. This chemical environment may have stabilized K-feldspar and thus restricted the albition process.

Albite formation as a replacement of plagioclase starts when the temperature is higher than about 75 °C (Boles and Ramseyer, 1988; Morad et al., 1990), but the minimum temperature of plagioclase albition is still poorly constrained. The albition of K-feldspar is commonly associated with illitization of kaolinite at greater burial depths and temperatures above about 125 °C (Morad et al., 1990). The calcite-cemented intervals in the studied sandstones reveal evidence of fibrous illite formation related to kaolinite. This indicate that the sediments may have been subjected to a high temperature event transforming kaolinite to illite. However, illite may also be formed from smectite at temperatures above about 60–70 °C (Hower et al., 1976; Bjørlykke and Jahren, 2015). Morphologically diagenetic illite show a lath-shaped texture that resembles the precursor smectite (Bauer et al., 2000). Fibrous illite on the other hand, is normally only found at high temperatures
Inclusions in authigenic cements of outcrop samples. Whereas $^{14}C$T plot, the true population has a vitrinite reflectivity population and a secondary population (population of vitrinite $(\text{indigenous population of vitrinite})$ and a secondary population (population of vitrinite $(\text{primary population})$ comprises 20 measurements while the secondary population (higher reflectivity population) comprises 20 measurements. There was little morphological difference between the particles from the two populations, but the true population material occurred as particles in isolation. (B) Frequency distribution histogram of homogenization temperatures $(T_h)$ for fluid inclusions in authigenic cements of outcrop samples. $T_h$ for the aqueous inclusions found in quartz cement varies between 89.8 °C and 128.6 °C with a mean value of 109 °C whereas $T_h$ for the aqueous inclusions in calcite cement varies between 100 °C and 138 °C with a mean value of 123 °C. (C) Plots of the equilibrium oxygen isotope fractionation between calcite and water as a function of temperature. The Z-value calculations suggest precipitation of calcite from meteoric pore waters. The Z-value calculations were based on Keith and Weber (1964). The gray shaded region illustrates the possible ranges of precipitation temperatures if the waters involved were of $(>120 \, ^\circ\text{C})$, as the energy barrier of nucleating these crystals are high on e.g., kaolinite (Wilkinson and Haszeldine, 2002; Lander and Bonnell, 2010). Fibrous illite in this study was only found locally together with the other evidence of hydrothermal alteration, further pointing to alteration along localized features (e.g., flow baffle). This suggests that the calcite-cemented sedimentary units of DG and WS have been affected by processes deviating from the normal burial diagenesis trend in the study area.

5.3. Mechanism of the hydrothermal induced diagenesis

Abnormally high temperature authigenic phases were distinguished in the calcite-cemented intervals both above and below the sill intrusion at distances up to more than five times the thickness of the sill. The conductive mechanism commonly used to explain heat transport away from sill intrusions fail to explain this observed diagenetic pattern. Instead, an explanation invoking that the formation of thermal convection hydrothermal cells can take place in a sedimentary basins consisting of interbedded highly porous-permeable sediments and semi-permeable or impermeable layers such as shale or calcite-cemented units (Genthon et al., 1990). Such cells may form focused flows along the flow baffle, limiting the high temperature alteration to these narrow zones.

Magmatic sills intruding into reservoirs rocks may change the petro-physical properties of the reservoir rocks intruded. Mobilization of hydrothermal fluids setting up convective fluid flow cells may affect strata located a considerable distance away from magmatic activity. The fluid flow will follow the most permeable sandstone strata or along permeable faults or fracture zones (Wilson et al., 2007).

There are several indications that calcite cemented intervals of the sediments have been subjected to higher temperatures. This is shown by the sericitized feldspars, albitized feldspars, carbonate fluid inclusions, and fibrous illite formation associated with kaolinite and relict feldspar, all pointing to hydrothermal alteration $(T > 120–140 \, ^\circ\text{C})$. The sericite (illite) could likely be detrital in origin but its nonexistence in the uncemented intervals suggests in situ formation related to hydrothermal fluids.

The source of the heat is presumably the magmatic sill intrusions penetrating the sediments at several levels. The magmatic sill intrusions are generally not sufficiently thick to thermally affect the entire sequence because the thickness of the sedimentary strata that could be affected by thermal heat generated by the sill intrusions due to conduction according to most estimates is approximately twice the width of the sill (Dow, 1977; Peters et al., 1978; Karlsen et al., 1998; Brekke et al., 2014). However, intrusion into shallow highly porous sediments can create hydrothermal convection cells, which is the most effective way of dissipating the excess energy (Einsle et al., 1980).

Fluid flow due to sill emplacement will most likely initiate close to the top of the gently dipping sill surface and also likely under the sill. This will be at either tip of the sill since sills tend to be saucer shaped structures (Jamtveit et al., 2004) and under the sill due to build-up of high fluid pressures sufficient to trigger fluid mobilization. The prevalent geometry of the igneous intrusions emplaced during the early Cretaceous has been identified as saucer-shaped in central Spitsbergen (Senger et al., 2013). This would explain why the sandstone strata has been altered only in areas where meteoric origin $(\delta^{18}O_{\text{vsmow}}$ ranges between $-3.5 \%$ to $-7.5 \%$). The isotopic composition of calcite is illustrated as contours. Calcite temperatures were calculated using Friedman and O’Neil (1977) the calcite-water fractionation factors equation $(1000 \ln (\text{calc-water}) = -2.78 \times 10^3 T^{-2.89})$ by assuming the water isotopic composition relative to SMOW based on the Z-value computation.
permeability differences resulting from existing carbonate cemented layers has led to channeling of the hydrothermal fluids. The heat perturbation related to the emplacement of relatively small sill bodies is short-lived (Galland et al., 2006; Parnell, 2010). However, the reactivity of the pore water will be highly enhanced when mixed with heated hydrothermal fluids because this may lead to undersaturation with respect to carbonates. This will then induce dissolution followed by precipitation of the carbonates whose dissolution-precipitation kinetics is known to be fast. As mentioned above, the diagenetic fingerprints of most of the highly permeable sedimentary succession and the vitrinite reflectance indicate that the sandstones have not been subjected to higher temperature. Only the low permeable calcite cemented intervals that have been exposed to the hydrothermal pore water flow, contain high temperature diagenetic phases. Based on the above observation, we proposed that buoyant fluids have been partly following tight carbonate cemented flow baffles, before migrating into the sedimentary strata (Fig. 13A). On a local scale, the amount of calcite cement in each cemented intervals vary (Fig. 13B). The observations of hydrothermal induced diagenesis associated with the cemented sedimentary strata both close (~1 m) and far away (~65 m) from the sill suggest hydrothermal fluid mobility through the strata (Fig. 13).

The main effect of the hydrothermal fluids is recrystallization of already existing carbonates, and localized formation of hydrothermal albite, illite and sericite. Most likely, the carbonate containing layers lost their remaining porosity due to the hydrothermal activity but these layers must already have been flow baffles before the magmatic intrusions were emplaced. Identification of fluid pathways both vertical and lateral has not been done in the investigated area. Fluid movement resulting from sill emplacements should be studied further in the area in order to elucidate possible flow patterns generated by sills better.

5.4. Implications of sill intrusions on reservoir quality

The hydrothermal fluids injected into cooler host rocks due to sill intrusion emplacement may have impacts on porosity

<table>
<thead>
<tr>
<th>Sample</th>
<th>$\delta^{13}C_{\text{V-PDB, Calcite}}$</th>
<th>$\delta^{18}O_{\text{V-PDB, Calcite}}$</th>
<th>Z-values</th>
</tr>
</thead>
<tbody>
<tr>
<td>DF-3</td>
<td>-7.5</td>
<td>-13.3</td>
<td>105</td>
</tr>
<tr>
<td>DF-4</td>
<td>-10.2</td>
<td>-19.1</td>
<td>97</td>
</tr>
<tr>
<td>WS-3</td>
<td>-9.7</td>
<td>-20.6</td>
<td>97</td>
</tr>
<tr>
<td>WS-4</td>
<td>-8.8</td>
<td>-18.8</td>
<td>100</td>
</tr>
</tbody>
</table>

Table 1. $\delta^{13}C$ and $\delta^{18}O$ pore-filling calcite cement stable isotope analyses for samples of upper Triassic to Lower Jurassic sediments from Wilhelmøya, Svalbard.
The data presented above show that diagenetic reactions related to hydrothermal fluid flow related to sill emplacement were observed in carbonate-cemented intervals only. No apparent diagenetic effects like porosity decrease related to the sill emplacement was observed within intervals lacking carbonate cements. This is in accordance with earlier findings where the total porosity even as close as a few centimeters from the sill intrusion was unaffected by contact diagenesis (Mckinley et al., 2001; Grove, 2014; Grove et al., 2017). However, the hydrothermal fluid flow set up by the sill intrusion affected already existing flow barriers or baffles related to carbonate cemented layers. This lead to recrystallization and possibly increased cementation along already cemented intervals resulting in increased reservoir compartmentalization. Increased compartmentalization due to recrystallization should be considered since thin carbonate cemented layers and/or shale layers might exist in many siliciclastic reservoir rocks. This observation can be extended to any sedimentary basin where interbedded sandstone with thin carbonate or shale layers is common. Porosity will probably be reduced within the cemented layers after recrystallization while the reservoir quality outside the recrystallized zones would probably not be affected significantly.

6. Conclusion

Investigating impact of igneous intrusions on reservoir properties is important in order to evaluate their impact on the hydrocarbon potential of the northwestern Barents Sea and sedimentary basins elsewhere. In this study, the influence of sill intrusions on diagenesis and hence reservoir quality was evaluated. Diagenetic evidences show that the upper Triassic and lower Jurassic sandstones at Wilhelmøya may be divided into intervals with different thermal histories; one being the bulk sediment, being largely unconsolidated and with a maximum burial temperature much lower than about 60–70 °C; and the second being thinner intervals that have experienced higher temperatures related to hydrothermal activity. The hydrothermally altered intervals are all tightly carbonate cemented layers. These intervals were most likely already carbonate cemented before the hydrothermal activity commenced, and would therefore have been flow baffles during the hydrothermal activity and thereby partly controlling the migration pathways of the buoyant hot fluids.

The hydrothermal fluid flow set up by the magmatic sill intrusions have affected the carbonate cemented sandstone intervals both close to the sill (∼1 m) and away (∼65 m) from the sill intrusion. The carbonate cemented intervals revealed high temperature hydrothermal induced reactions such as recrystallization of carbonate cements, localized sericitization of feldspars, albitionization of both feldspar and plagioclase, and formation of fibrous illite nucleated on kaolinite. Within the intervals not affected by hydrothermal activity, there was no indication of hydrothermal induced diagenetic changes. This implies that the sill intrusion emplacement has not affected the porosity of these intervals.

Most of the available literature has focused on the effect of sill intrusion through conduction only. This work shows that igneous
sill intrusion can also affect host rock intervals due to heat transfer through fluid flow. Possible hydrothermal fluid convection cells resulting from emplacement of sills should therefore be assessed together with conduction heat transfer when the influence of sill intrusions on reservoir quality is evaluated. The results from this study are applicable to the more general case of sedimentary basins having equivalent settings to the Wilhelmsøya sediments.

Performing numerical modeling regarding the hydrothermal convection cell is beyond the scope of this study; however, the results obtained from this study could likely serve as inputs to perform such type of modeling in a future study. This will allow performing mass and energy balance calculations in order to understand the interplay between porous and permeable sediments and magmatic intrusions. This may enhance our quantitative predictive ability regarding reservoir quality evolution in such settings.

Acknowledgements

This work has been (partially) funded by the project “Reconstructing the Triassic Northern Barents shelf; basin infill patterns controlled by gentle sags and faults” (Trias North--www.mn.uio.no/triasnorth/) under grant 234152 from the Research Council of Norway and with financial support from Tullow Oil Norge, Lundin Norway, Statoil Petroleum, Edison Norge and RWE Dea Norge. We would like to thank reviewer Clayton Grove and anonymous reviewer, for constructive and insightful comments, which significantly improved an earlier version of the manuscript. Associate Editor Dr. Nick Roberts is also kindly thanked for editorial handling of the manuscript by giving his invaluable time. We are immensely thankful to Knut Bjørlykke for sharing his wealth of wisdom in diagenesis with us during the course of this research. We thank Maarten Aerts and Berit Løken Berg for guidance and assistance during the XRD and SEM work, respectively. We are also grateful to Kim Senger and Mark Mulrooney from Department of Arctic geology, University Centre in Svalbard, Norway and Kei Ogata from Vrije University, Amsterdam, Holland, for sharing their knowledge regarding the effect of sill intrusion emplacement on fluid flow in country rocks. Moreover, Mark Mulrooney and Kei Ogata are thanked for providing us quality outcrop pictures and for assistance during fieldwork. We would also like to thank Yingchang Cao from University of petroleum, Qingdao, China for allowing us to use their facility to perform fluid inclusion analyses on quartz and carbonate cement.

Appendix A. Supplementary data

Supplementary data related to this article can be found at https://doi.org/10.1016/j.gsf.2018.02.015.

References
